

IDENTIFICATION OF A POSSIBLE SEISMIC GAP NEAR
UNALASKA ISLAND, EASTERN ALEUTIANS, ALASKA

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Abstract. A portion of the eastern Aleutians, from about 168°W and 160°W, appears to be a tectonic transition between the oceanic arc structure to the west and the continental arc structure to the east. The 1200 km long aftershock zone of the 1957 Andreanof-Fox Islands earthquake, $M_w = 9.1$, extends 700 km into the eastern portion of this zone. This 200 km long segment, which is located near Unalaska Island, shows deformation quite different from that of the remaining 1000 km long zone. It experienced aftershocks along only a narrow (20 km width) zone to the arcward (northern) edge, in remarkable contrast to the broad distribution of aftershocks to the west. Since 1957, the interior of this segment has produced only two earthquakes of magnitude 5 or greater. Prior to the 1957 main shock, one near the western end of the entire aftershock zone (168°W), the other, a more dense cluster, occurred near the western edge of the Unalaska segment (163°W). Travel times of the 1957 tsunami to tide gauges in western North and South America indicate the eastern extent of the tsunami generating area was situated at or near the western boundary of the Unalaska segment. If the Unalaska segment ruptured in 1957 it must either have undergone a complete rupture, so that seismic and tsunami energy were concealed in the coda of the main shock, or have slipped in a rupture process so that it was not an efficient tsunami source. The latter possibility clearly exists, however, since the Unalaska segment did not rupture in 1957. If it is a seismic gap, and considering its high potential for producing an earthquake of large magnitude, this possibility has significant implications for the evaluation of seismic and tsunami hazards in the eastern Aleutians.

by L.S. House
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Introduction

The concept of seismic gaps has been highly successful in identifying those portions of simple plate boundaries that are likely sites of future large ($M > 7$) earthquakes [Fedotov, 1965; Mogi, 1968a; Sykes, 1971; Kelleher et al., 1973]. Evaluating the seismic potential of segments of a plate boundary represents a refinement of the seismic gap concept and provides the impetus to focus earthquake prediction studies on areas that seem most likely to produce large earthquakes in the near future (one to several decades) [McCann et al., 1977]. We consider an area to be a seismic gap if it is part of an active plate boundary and has not experienced a large earthquake in the past 30 years. In order to assign a high seismic potential to a gap, they add the additional requirement that the area must have broken in a large earthquake at least once in the historic past.

To identify a portion of a plate margin as a seismic gap requires delineating the rupture areas of previous large earthquakes along it. This is usually done by assuming that the aftershock area coincides with the rupture area of an earthquake. This assumption is supported by studies demonstrating that aftershock areas of great ($M > 7.8$) earthquakes tend to abut without significant overlap [e.g., Fedotov, 1965; Mogi, 1968a; Kelleher, 1970; Sykes, 1971]. Felt areas of intensity VIII-IX are also used to define rupture areas [Kelleher, 1972] if aftershock data are inadequate.

Aftershock data from an earthquake in 1957 that broke a large portion of the Aleutian plate margin show an unusual distribution. This earthquake is one of the largest events ever recorded, with a magnitude, M_w , of 9.1. Its aftershock zone stretches 1200 km along the central and eastern Aleutians, from 180°W to 163°W. In the western 1000 km of this zone, the main segment, aftershocks distribute over a zone about 80 km wide, measured normal to the arc. In contrast, aftershocks in the eastern 200 km segment, the Unalaska segment, define a narrow zone less than 20 km

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EASTERN ALBERTA, ALASKA

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IDENTIFICATION OF A POSSIBLE SEISMIC GAP NEAR
UNALASKA ISLAND, EASTERN ALEUTIANS, ALASKA

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Abstract. A portion of the eastern Aleutians, between about 168°W and 160°W, appears to be a major tectonic transition between the oceanic arc structure to the west and the continental arc structure to the east. The 1200 km long aftershock zone of the 1957 Andreanof-Fox Islands earthquake, $M_w = 9.1$, extends 200 km into the western portion of this zone. This 200 km long segment, which is located near Unalaska Island, underwent deformation quite different from that of the remaining 1000 km long zone. It experienced aftershocks along only a narrow (20 km width) zone at its arcward (northern) edge, in remarkable contrast to the 80 km width of the aftershock zone to the west. Since 1957, the interior of this segment has produced only two earthquakes of magnitude 5 or larger. Two clusters of events occurred prior to the 1957 main shock, one near the western end of the entire aftershock zone (180°W), the other, a more dense cluster, occurred near the western edge of the Unalaska segment (168°W). Travel times of the 1957 tsunami to tide gauges in western North and South America indicate that the eastern extent of the tsunami generating area was situated at or near the western boundary of the Unalaska segment. If the Unalaska segment slipped in 1957 it must either have undergone delayed rupture, so that seismic and tsunami energy were concealed in the coda of the main shock, or have slipped in a rupture process so slow as to be an inefficient tsunami source. Another possibility clearly exists, however, namely that the Unalaska segment did not rupture in 1957. If it did not, the Unalaska segment is a seismic gap, and could also have a high potential for producing an earthquake as large as magnitude 8. This possibility has serious implications for the evaluation of seismic and tsunami hazards in the eastern Aleutians.

Introduction

The concept of seismic gaps has been highly successful in identifying those portions of simple plate boundaries that are likely sites of future large ($M \geq 7$) earthquakes [Fedotov, 1965; Mogi, 1968a; Sykes, 1971; Kelleher et al., 1973]. Evaluating the seismic potential of segments of a plate boundary represents a refinement of the seismic gap concept and provides the impetus to focus earthquake prediction studies on areas that seem most likely to produce large earthquakes in the near future (one to several decades) [McCann et al., 1979]. McCann et al. [1979] consider an area to be a seismic gap if it is part of an active plate boundary and has not experienced a large earthquake in the past 30 years. In order to assign a high seismic potential to a gap, they add the additional requirement that the area must have broken in a large earthquake at least once in the historic past.

To identify a portion of a plate margin as a seismic gap requires delineating the rupture areas of previous large earthquakes along it. This is usually done by assuming that the aftershock area coincides with the rupture area of an earthquake. This assumption is supported by studies demonstrating that aftershock areas of great ($M > 7.8$) earthquakes tend to abut without significant overlap [e.g. Fedotov, 1965; Mogi, 1968a; Kelleher, 1970; Sykes, 1971]. Felt areas of intensity VIII-IX are also used to define rupture areas [Kelleher, 1972] if aftershock data are inadequate.

Aftershock data from an earthquake in 1957 that broke a large portion of the Aleutian plate margin show an unusual distribution. This earthquake is one of the largest events ever recorded, with a magnitude, M_w , of 9.1. Its aftershock zone stretches 1200 km along the central and eastern Aleutians, from 180°W to 163°W. In the western 1000 km of this zone, the main segment, aftershocks distribute over a zone about 80 km wide, measured normal to the arc. In contrast, aftershocks in the eastern 200 km segment, the Unalaska segment, define a narrow zone less than 20 km

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wide. This narrow zone of activity extends the northern edge of the aftershock zone defined by the main aftershock segment. To better understand the behavior of the Unalaska segment of the 1957 aftershock zone, we studied earthquake data in the area of the 1957 earthquake from the past 33 years and tide gauge recordings of the tsunami excited by the 1957 earthquake.

We find that the Unalaska segment, if it underwent deformation in 1957, did so in a manner quite different from the manner of the main segment of the 1957 aftershock zone. Either it ruptured in a delayed event, whose seismic and tsunami signatures were masked in the coda of the main rupture or, alternatively, it ruptured with a very slow source process. The seismic energy of a very slow (longer than several tens of minutes) rupture process in the Unalaska segment might not be distinguishable from the coda of the main rupture and would be inefficient in generating a tsunami.

A third possibility also exists, however. The Unalaska segment may not have ruptured at all in 1957. If it did not, it cannot have broken since about 1902 [Sykes et al., 1981] and would therefore be a seismic gap. It could also have a high seismic potential [McCann et al., 1979] and would produce an earthquake as large as magnitude 8 if it were to rupture completely in a single earthquake. Thus, the evaluation of seismic and tsunami hazards in the eastern Aleutians could be substantially affected if the Unalaska segment did not rupture in 1957.

Figure 1 (top) shows the location of the 1957 earthquake rupture zone and its relation to the rest of the Alaska-Aleutian plate margin. It also shows the locations of three seismic gaps identified by previous studies [Kelleher, 1970; Sykes, 1971; Davies et al., 1981]. From the west, these are: the western Aleutians, an area where relative motion is highly oblique [Sykes, 1971]; the Shumagin Gap, discussed by Davies et al. [1981]; the Yakutat Gap, described by McCann et al. [1980] and Lahr et al. [1980]. The possible Unalaska Gap is the area at the eastern end of the 1957 aftershock zone, between about 164°W and 167°W, labelled as a queried gap.

Tectonic Setting

Geologic evidence indicates that the region of the Alaska-Aleutian arc between 168°W and 160°W is a transitional region between an oceanic type arc structure to the west and a thicker, more continental type arc structure to the east. The Bering Sea Shelf, indicated by the 100 fathom bathymetric contour in Figure 1 (bottom) is believed to be an extinct plate margin that was active during the Mesozoic [Marlow et al., 1977]. At that time, the active plate margin trended subparallel to the present margin westward to about 162°W then began to curve northwestward to follow the trend of the Bering Sea Shelf [Moore, 1972; Scholl et al., 1975; Moore and Connolly, 1977]. The active plate margin jumped to its

present position along the Aleutian arc prior to the early Tertiary [Scholl et al., 1975]. The trend of the Bering Sea shelf edge intersects the Aleutian arc at about 165°W, near the eastern edge of the Unalaska segment.

Geophysical data also indicate that the Unalaska region is a transitional region along the Alaska-Aleutian arc. The volcano-trench separation is nearly constant at about 160 km along the entire portion of the Aleutians from Kiska Island (about 178°E) to Akutan Island, at about 166°W [Davies, 1975]. To the east of Akutan Island, it systematically and smoothly widens, until at Augustine Volcano, in Cook Inlet (about 154°W) it measures nearly 400 km. The Aleutian Terrace, a flat, well defined forearc basin about 70 km wide, extends along the entire central and eastern Aleutians east of about 180°W, but begins to narrow near Akutan Island and has disappeared entirely by 160°W [Nishenko and McCann, 1979]. Similarly, the trend of a prominent 100-150 mgal gravity high, which follows the volcanic line in the central and eastern Aleutians, diverges from this trend at Akutan Island where it turns eastward and follows the trend of the shelf edge break [Watts et al., 1976].

The transitional nature of the Alaska-Aleutian arc between about Unalaska and Unimak Islands (for locations, see Figure 4) may influence the rupture process of earthquakes whose ruptures approach or extend into it. The unusual behavior of the Unalaska segment during and since the 1957 earthquake may be an example that merits special attention.

Seismicity Data

Mogi [1968b], Kelleher [1970] and Sykes [1971] delineated the rupture zone of the 1957 earthquake from its aftershocks, which extend 1200 km eastward along the central and eastern Aleutians from about 180°W. This is the longest aftershock zone ever identified. The earthquake involved bilateral rupture, since the epicenter of the main shock at 51.6°N, 175.4°W, is located well within the aftershock zone (Figure 2). Good quality long period seismograms of this event are sparse [H. Kanamori, written communication, 1979] so its moment, 6×10^{29} dyne cm, and magnitude, $M_w = 9.1$, are estimated from the aftershock area [Kanamori, 1977].

The bottom half of Figure 1 [after Sykes, 1971] shows, as large X's, well-located aftershocks of the 1957 earthquake that occurred during March 1957 (about 20 days) and includes shocks of all magnitudes. Several features of the aftershock distribution are notable. First, along most of the length of the aftershock zone, aftershocks are scattered between the island arc and the Aleutian Trench. This is the main segment of the aftershock zone. Aftershocks in the eastern 200 km, the Unalaska segment, occur only along the northern (arcward) edge of the aftershock zone. Second, a large number of epicenters plot within

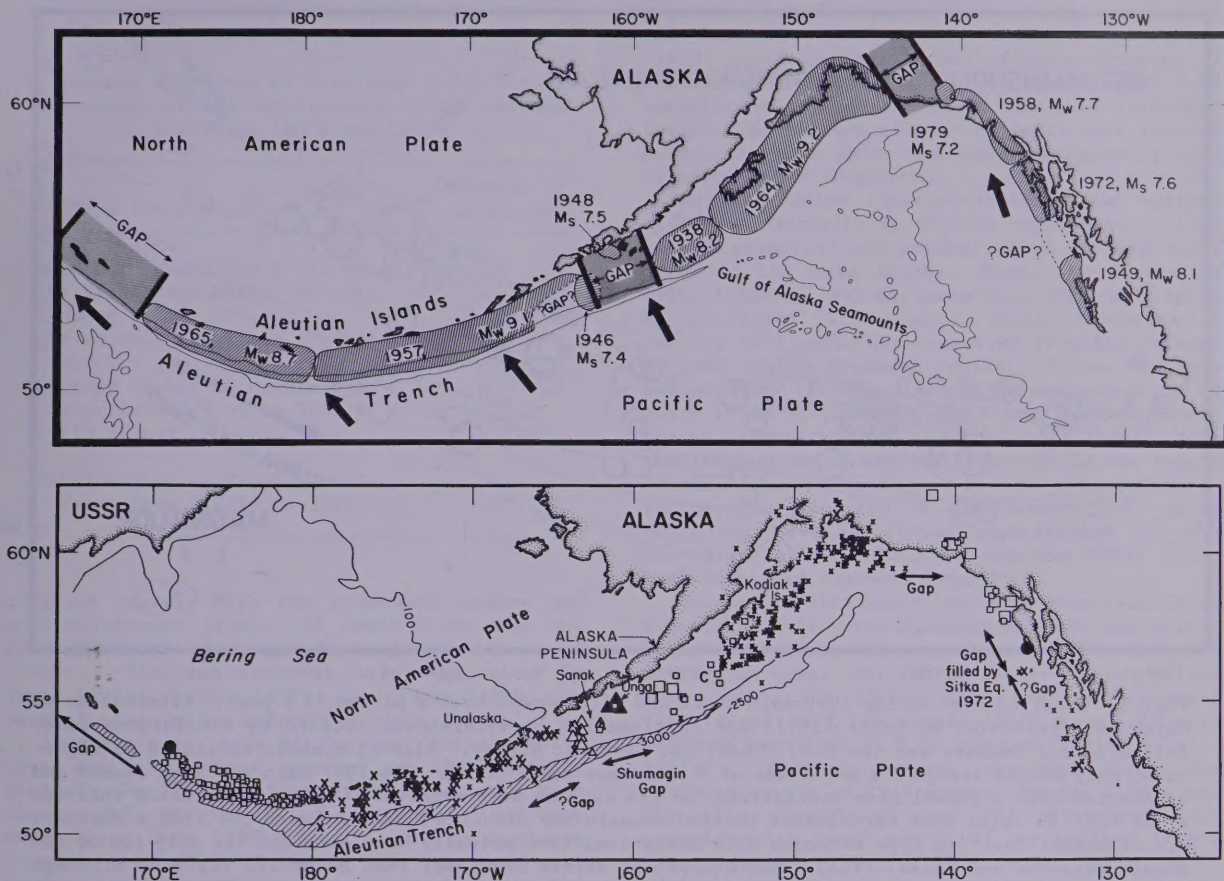


Fig. 1. Top: Location map and identification of aftershock zones of major earthquakes and previously identified seismic gaps in Alaska and the Aleutians. The possible Unalaska Gap is the area labelled as a queried gap near 165°W. Note the proximity of the 1946 tsunamigenic earthquake to the possible Unalaska Gap, and also that the Shumagin Gap nearly abuts the eastern edge of the possible Unalaska Gap. Arrows indicate direction of relative convergence. [After Davies et al., 1981]. Bottom: Map of relocated aftershocks of recent major earthquakes in Alaska and the Aleutians. Only those aftershocks of the 1957 earthquake that occurred in March of 1957 are included; these are plotted as large X's. Note how far trenchward the aftershocks extend between 180°W and 166°W as compared to the very narrow band between 166°W and 163°W. Also note that 3 aftershocks plot within the aftershock zone of the 1946 earthquake, which is identified by triangles (see text for discussion of these events). Bathymetry is in fathoms (1 fathom = 1.83 m). [After Sykes, 1971].

the Aleutian Trench. We shall demonstrate that these events resulted from normal faulting triggered or stimulated by the large underthrust event, and are not aftershocks in the strict sense of defining the rupture zone. Third, several aftershocks of the 1957 earthquake plot within the aftershock zone of the 1946 tsunamigenic earthquake (identified by triangles in Figure 1). No magnitudes are assigned to these events in the International Seismological Summary for 1957, but from the small number of stations reporting, they are likely to be magnitude 6 or smaller.

A more detailed plot of aftershocks during the

first year after the 1957 earthquake is shown in Figure 2 and contains only shallow (depth ≤ 60 km) earthquakes with assigned magnitude of 5.0 and larger. Epicenters of these events are from the relocations of Sykes [1971] and from the locations of the International Seismological Summary and the U.S. Coast and Geodetic Survey. From a frequency-magnitude plot, we estimate that the data set is complete only for earthquakes larger than about magnitude 6 1/4. The distribution of seismicity in this figure is somewhat more heterogeneous than that plotted in Figure 1, with much clustering evident. Within

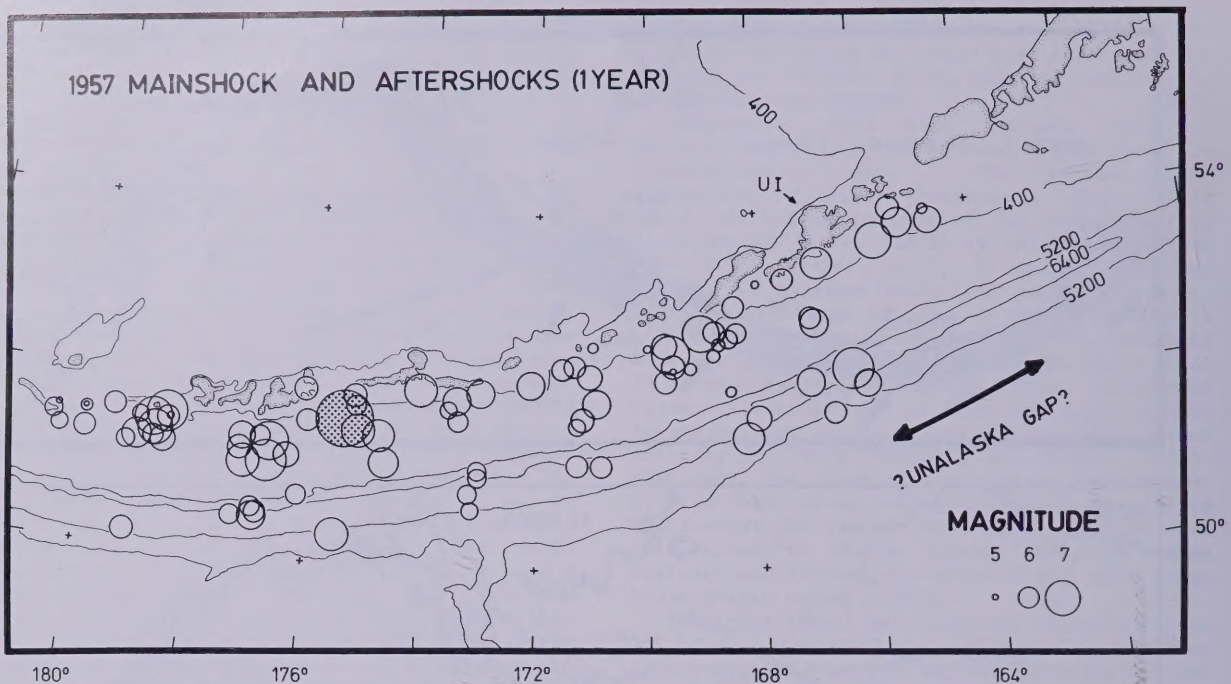


Fig. 2. Detailed plot of the 1957 main shock and aftershocks over a period of 1 year. Aftershocks of March 1957 relocated by Sykes [1971] are supplemented by aftershocks located by the International Seismological Summary and the U.S. Coast and Geodetic Survey. Size of symbols is scaled to magnitude; only events assigned a magnitude of 5 or larger are plotted. The 1957 main shock is shaded and is plotted with a symbol size appropriate for its surface wave magnitude (M_s) of 8.2, rather than its M_w of 9.1. We infer that earthquakes located beneath the Aleutian Trench between the 5200 m contours are of normal faulting type although such mechanisms have actually been obtained for only two of the events [Stauder and Udias, 1963]. Earthquakes at depths shallower than 80 km are included although none occurred that were deeper than 60 km. UI indicates the location of Unalaska Island. Bathymetry is in meters.

the main portion of the aftershock zone, events scatter over a width normal to the arc of about 80 km. Note the large number of aftershocks with magnitudes about 7. For comparison, no aftershocks larger than magnitude 6.5 occurred in the aftershock zone of another great earthquake nearby, the 1964 Alaska earthquake [Davies et al., 1981] (Figure 1).

Activity within the Unalaska segment of the aftershock zone defines an eastward extension of the northern margin of activity to the west of about 167°W longitude. Depths of these 5 events, based on pP-P times reported in the International Seismological Summary, range from 45 to 60 km. This depth range is appropriate for the deeper, or down dip, edge of the main thrust zone or seismic portion of the plate interface in the Aleutians [Davies and House, 1979]. Stauder and Udias [1963] obtained an underthrust-type focal mechanism for the westernmost event of the cluster of five (53.6°N, 165.8°W, $M = 7$). Although they are not well constrained, both nodal planes dip steeply and strike subparallel to the trend of the

arc. It is plausible, therefore, that the aftershock activity located at the northern edge of the Unalaska segment represents underthrust-type motion along the deeper edge of the plate interface. The remaining shallower portion of the interface was aseismic, at least for earthquakes larger than about magnitude 6 1/4.

Another characteristic of the aftershock activity is the large number of events whose epicenters plot within the Aleutian Trench (Figure 2). Focal mechanisms of two of these events [Stauder and Udias, 1963] show normal faulting, which is a typical focal mechanism for earthquakes beneath trenches. During the 15 years from 1959 to 1974 (see Table 1), earthquakes of magnitude 6 and larger occurred beneath the trench in areas offshore of the 1957 earthquake on an average of less than once every 2 years. None were located within this area during the 12 years prior to the main shock. Fully 13 events occurred in this area during the first year after the 1957 main shock and a total of 19 occurred there in the 2 years following the main shock. These numbers

Table 1. Normal Faulting Following the
1957 Earthquake

Numbers of events of magnitude ≥ 6.0 with
epicenters within the Aleutian Trench between
longitude 160°W and 180°W

Time Interval	Number of Events
Before the main shock (12 years)	0
The first year after the main shock	13
The second year after the main shock	6
Average number of events per year from 3/9/59 to 3/9/74 (15 years)	<u>0.40</u>
Total Number of Events Consi- dered from 3/9/57 to 3/9/74	25

contrast sharply with the quiescence before and near quiescence after. It seems clear that the 1957 mainshock stimulated these large numbers of events. Similar phenomena have been noted by Sykes [1971], Spence [1977], and Hanks [1979] for several other major underthrust-type events.

Interestingly, none of the many normal fault-type events occurred seaward of the possible Unalaska Gap in the first year after the main shock. Only one did in the second year, and none have since then (1959-1979). If the Unalaska segment did not rupture in 1957 we would not expect induced activity to occur beneath the adjacent portion of the trench. The distribution of such activity within the rest of the 1957 rupture zone is quite heterogeneous, however, so the lack of it seaward of the Unalaska segment may be only fortuitous.

Figure 3 illustrates earthquake activity during 11 years before the 1957 main shock. This figure clearly shows two distinct clusters. One is at the western end of the aftershock zone, and developed over about 3 years before the main shock. The other plots at the eastern end of the main portion of the aftershock zone (compare Figures 2 and 3). The latter cluster primarily developed as a swarm over a period of about 1 week in early January 1957, although it also includes several events from as early as 1950. Mogi [1968b] noted the occurrence and location of the swarm of January 1957 and suggested that it represented an area weakened by partial failure prior to the main rupture. If the area of the 1957 aftershock zone had been identified as a seismic gap prior to 1957, the occurrence of these two very clear earthquake clusters might have prompted the issuance of an intermediate-term earthquake prediction for the area.

The two clusters seem highly suggestive of

pre main shock failure processes at the margins of the eventual rupture zone. Kelleher and Savino [1975] noted the occurrence of such clusters of activity prior to several great (magnitude ≥ 7.8) underthrust type earthquakes. By this interpretation, the eastern cluster suggests that rupture during the main shock may not have extended east of about 168°W longitude.

An alternative interpretation of the eastern cluster of activity is that it represents failure of a distinct stress concentration (or asperity) prior to the main shock. Thus, the events in 1950, 1952 and 1955 may have resulted from a build up of stress in the area, while the swarm of January 1957 could have resulted from the failure of that highly stressed region. Since the area had already failed prior to the main rupture of March 1957, it probably would not produce aftershocks associated with the March main shock. Note that in fact no aftershocks occurred in the region of this cluster in Figure 2. In this interpretation, the location of the January 1957 swarm would not have particular significance for the dimensions of the main 1957 rupture, which could have continued eastward beyond it.

Seismicity data since 1957 also show remarkable quiescence within the Unalaska segment compared to seismicity within the main portion of the 1957 aftershock zone. During the entire 21 year period from 1957 to 1979, the Unalaska segment experienced only two events (both $m_b = 5.0$) with assigned magnitude of 5 or greater [Davies et al., 1981]. Numerous events occurred along the northern margin of this segment during this time. The main segment of the 1957 aftershock zone experienced a large number of earthquakes larger than magnitude 5 during this time. These events could represent aftershock activity in a general sense. Thus, the Unalaska segment does not seem to have experienced any significant activity that could be interpreted as aftershocks in any sense.

Study of the Source Area of the 1957 Tsunami

Method. Large underthrust-type earthquakes commonly generate sizable tsunamis [Cox, 1963; Hatori, 1970; Ando, 1975; Abe, 1979; Nishenko and McCann, 1979], and the 1957 earthquake was no exception [Salsman, 1959; Abe, 1979; Nishenko and McCann, 1979]. The size of the tsunami source area generally corresponds well to the aftershock areas of large earthquakes [Hatori, 1970; Nishenko and McCann, 1979].

As noted by Salsman [1959] the tsunami from the 1957 earthquake was generated over a region of sizeable dimensions, rather than at a point. Tsunami energy excited by the eastern portion of the 1957 rupture zone would be recorded by tide gauges along the western coasts of North and South America. Therefore, the travel times to gauges in these locations would constrain the eastern extent of the tsunami generating area.

The speed of tsunami waves can be approximated quite closely by the simple formula: $S = (gd)^{1/2}$,

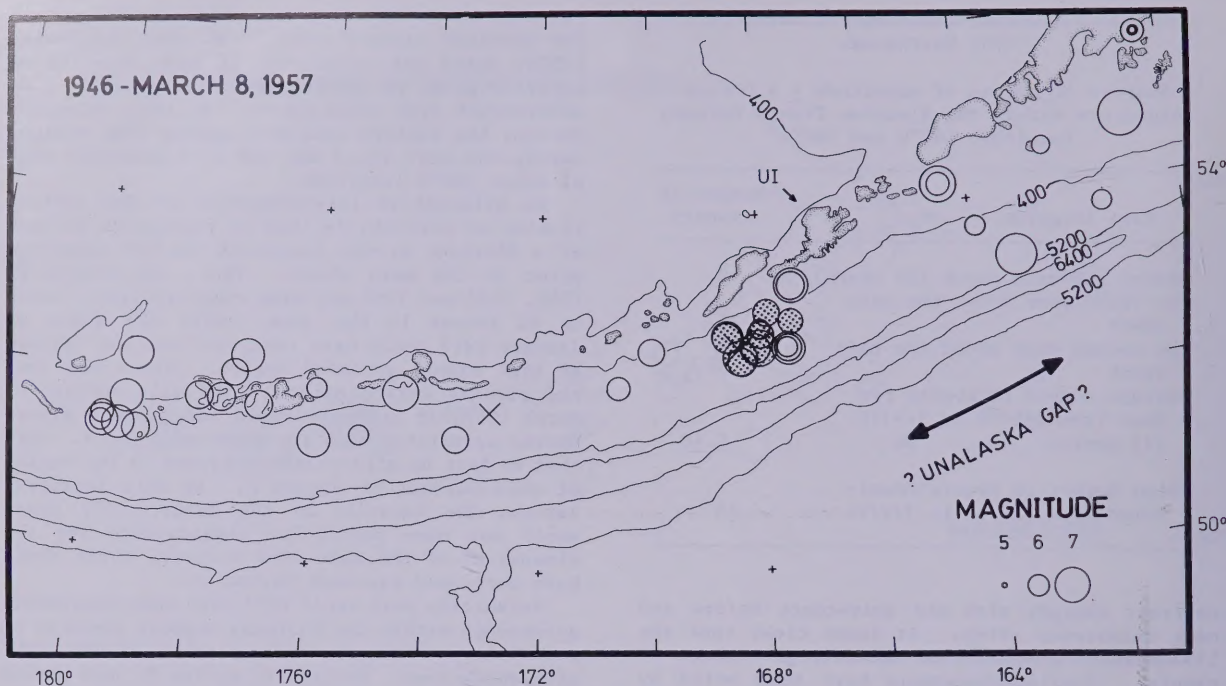


Fig. 3. Plot similar to Figure 2, but of earthquakes from 1946 to just before the 1957 main shock. Earthquakes at depths shallower than 60 km are plotted as circles; those at depths of 60 to 80 km as X's. Note the clustering of epicenters near 180°W and near 168°W. Earthquakes that are part of a swarm that occurred in early January 1957 are shaded. Note also the location of the 1946 tsunamigenic earthquake, immediately to the left of the 5200 meter label.

where g is the acceleration due to gravity ($\sim 9.8 \text{ m/s}^2$) and d is the water depth [Cox, 1963; Li and Lam, 1964]. From this formula, travel-time diagrams can be used to locate the extent of the source. This approach, however, yields estimates of the eastern extent of the source of the 1957 tsunami that are so scattered it is not possible to distinguish whether the source area extends eastward to 164°W or only extends to 168°W [L. R. Sykes, unpublished data, 1979]. One possible source of relatively large errors in this approach is the calculation of tsunami speed in the very shallow water portions of the path, where small depth errors can produce large travel time errors.

We use the relative location technique to locate the eastern extent of the source area of the 1957 tsunami more precisely. The nearby 1946 Aleutian earthquake provides a well-located tsunami source for comparison.

As noted by Green [1946] and Sheperd et al. [1950] the 1946 Aleutian earthquake generated an extremely large tsunami that was destructive over much of the Pacific Coast. This large tsunami is especially notable because of the moderate surface-wave magnitude of the earthquake ($M_s = 7.4$, Sykes [1971]; Kanamori [1972a]; Fukao [1979]). The aftershock zone of this event covers

an area of only about 100-150 km in diameter (Figures 1 and 4) that extends to the NE of the epicenter of the main shock at 53.3°N, 163.2°W [Sykes, 1971]. Green [1946] and Bodle [1946] studied the tsunami of this event and concluded that its source area was confined very near the epicenter. Davies et al. [1981] find that the aftershock and tsunami generating areas are nearly identical for the 1946 shock. Since many tide gauges recorded the tsunami of both this event and the 1957 earthquake, it is a good reference event.

Results. The results of the relative relocation are compiled in Table 2. Travel times of the 1957 tsunami to eastern Pacific stations are 18 to 27 minutes longer than those of the 1946 event to the same tide gauges. If the tsunami source of the 1957 event had extended over the entire aftershock zone indicated in Figure 1, the travel time differential for the two events would have been small, about 6 minutes or less.

Several assumptions were made in our analysis. These were: 1) the tsunamis of 1946 and 1957 followed nearly identical paths, except in the immediate area of the 1957 source; 2) the arrival times of the 1957 tsunami correspond to the first arriving tsunami energy; 3) rupture during the 1957 earthquake propagated smoothly between the hypocenter and the easternmost extent of the

Table 2. Comparison of Travel Times from the 1957 and 1946 Tsunamis

Tide Gauge	1957 Tsunami		1946 Tsunami		Δ (1957-1946) (min)	From Tsunami Source to Tide Gauge		Tsunami Speed Along Differential Path Between 1946 and 1957 Source Areas (km/min)	Distance West from 1946 Tsunami to 1957 Tsunami (km)
	Arrival time ¹	Travel time ¹	Arrival time ¹	Travel time ¹		Azimuth (°)	Distance (km)		
Neah Bay, Washington	19:20 ²	4:57	17:00 ³	4:31	+26	82°	3,000	14.2	327
Crescent City, California	19:30 ⁴	5:07	17:09 ⁴	4:40	+27	93°	3,400	13.6	327
San Francisco	20:18 ²	5:55	18:00 ³	5:31	+24	98°	3,789	13.6	283
La Jolla	20:57 ⁴	6:34	18:43 ⁴	6:14	+20	99°	4,500	13.6	226
Antofagasta	32:04 ²	17:41	29:50 ³	17:21	+20	100°	12,477	13.6	226
Valparaiso	32:48 ²	18:25	30:36 ³	18:07	+18	108°	13,165	13.5	210

¹Units are hour:minute.

²Published arrival time, Salsman [1959].

³Published arrival time, Green [1946].

⁴Reread from records published in Green [1946] or Salsman [1957]. Reread arrivals agreed with published arrivals within 3 minutes.

tsunami source area. Assumption 1 probably is quite good, since the distance between source and receiver is never less than 10 times the distance between the two sources. Assumptions 2 and 3 may not be valid; the discussion section considers them in more detail.

For the individual differential travel paths between the 1946 and 1957 tsunami source areas we calculate depths that are distance weighted averages of the depths along each travel path. We obtain average tsunami speeds (listed in Table 2) from the average depths by using the formula above, and reduce differential travel times of the two tsunamis to relative locations of the eastern extent of the 1957 tsunami source area compared to that of the 1946 tsunami. Our calculations corrected the differential travel times for the finite rupture velocity of the 1957 earthquake, which we chose as 3.5 km/s. Our results do not depend strongly on the actual velocity chosen since the rupture velocity of earthquakes generally is more than 10 times the velocity of tsunamis.

The last column of Table 2 lists the relative locations obtained, which are plotted in Figure 4. The solid lines in the figure are determinations from travel times reread from published records, dashed lines are from published travel times.

Uncertainties in the determination of the eastern extent of the 1957 tsunami arise

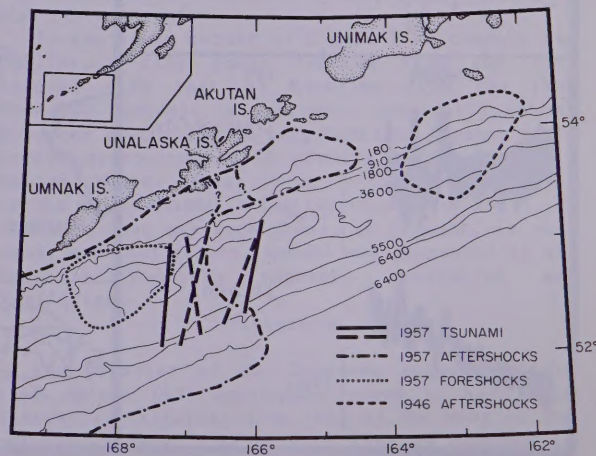


Fig. 4. Detailed map and summary of information about the eastern portion of the aftershock zone of the 1957 earthquake. The eastern extent of the 1957 tsunami is indicated by a solid line for stations that the authors could reread arrival times and by a dashed line when the determination is based on published arrival times (see text for sources). Other lines enclose the aftershock zones of the 1957 and 1946 earthquakes, and a swarm that occurred 2 months before the 1957 main shock. Bathymetry is in meters.

principally from two sources - error in reading the arrival times, which we estimate to be about ± 2 minutes, and error in the tsunami speeds along the differential travel path, which we estimate to be at most $\pm 20\%$. Error limits of 20% in the tsunami speed were chosen arbitrarily; actual errors seem unlikely to exceed this, since depth errors greater than about 25% would be required. The uncertainty resulting from errors in tsunami speed are ± 65 km or less, and dominate the uncertainty from travel time errors (about 30 km), so we take ± 65 km as a reasonable uncertainty in our estimates of the source location. This corresponds to about 1° of longitude in Figure 4. Individual estimates fall within this uncertainty and determine the eastern extent of the 1957 tsunami source at about 166.5°W to 167°W . This location is nearly identical to that of the eastern end of the main portion of the aftershock zone (see Figure 4).

We note also that the tsunami data excludes the possibility of significant tsunami generation within the aftershock zone at the northern edge of

the Unalaska segment. Consider, for example, if seismic rupture continued eastward into the Unalaska segment along a steeply dipping imbricate fault. Walcott [1978] inferred that the 1931 Hawke's Bay earthquake in New Zealand may have broken such a fault. This type of fault would outcrop substantially arcward of the outcrop of the more shallowly dipping fault zone that we presume ruptured within the main segment of the 1957 aftershock zone. A steeply dipping rupture geometry might produce the pattern of aftershocks observed within the Unalaska segment, as well as the focal mechanism of one of these [Stauder and Udias, 1963] which was described earlier. The travel times of a tsunami produced by this process, however, would be substantially shorter, by as much as 10 minutes or more, than the observed travel times.

Reports of near source tsunami effects, while not definitive, seem consistent with our identification of the source area. A published report that the 1957 tsunami reached 40 feet high (12 m) at Scotch Cap [Brazee and Cloud, 1959] at

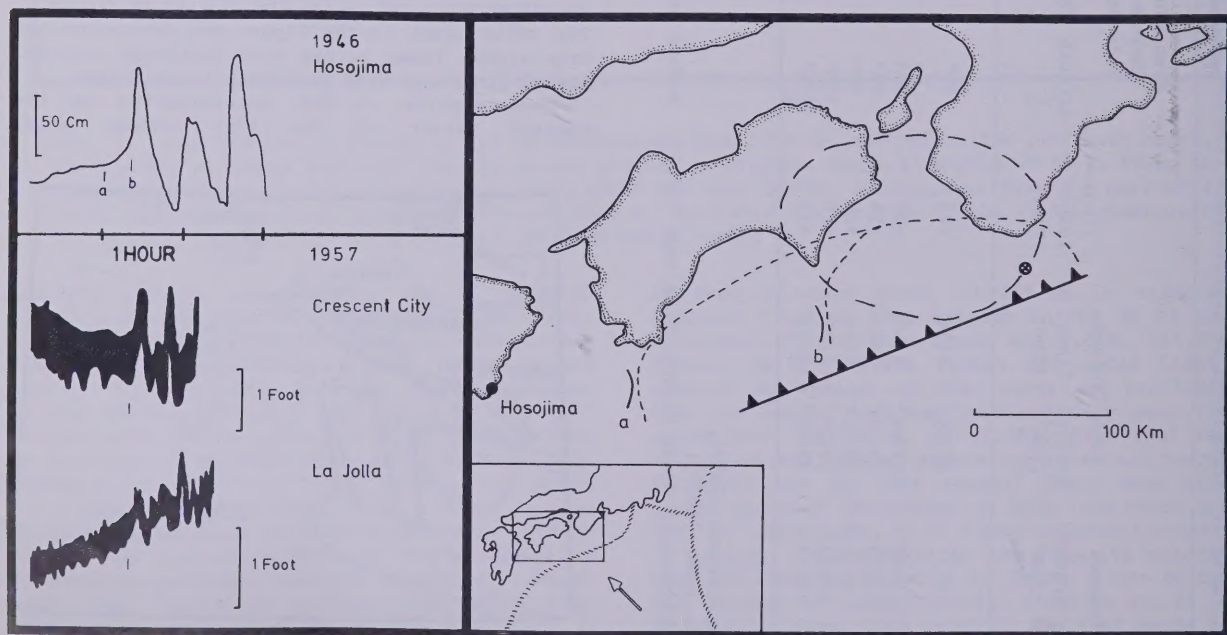


Fig. 5. Comparison of tide gauge recordings from the 1957 Aleutian Islands earthquake and the well-studied 1946 Nankaido earthquake, southwest Japan. Right: map of southwest Japan, showing the area of the 1946 Nankaido earthquake. The location of the main shock is indicated by a star and the aftershock zone after 1 month by a coarse dashed line. Hypothesized location of tsunami source area is indicated by the fine dashed line. Two arcs, labelled 'a' and 'b' are the locations of two phases identified on tide gauge record from Hosojima, to the southwest [after Ando, 1975]. Toothed line indicates surface trace of rupture zone obtained from geodetic data [after Fitch and Scholz, 1971]. Left: tracings of tsunami from 1946 Nankaido earthquake recorded at Hosojima (top) and of a tsunami from 1957 Aleutian earthquake recorded at Crescent City and La Jolla, California (below). Tracings are at the same time scale; the time between tick marks is 1 hour. Arrival times are indicated by tick marks under records; 'a' and 'b' are identifications of Ando [1975]. Tracings are from Ando [1975] and Salsman [1959].

the southwest end of Unimak Island (see Figure 4) might suggest that the source of such a large wave must be closer to Scotch Cap than our determination of the source is. For comparison, however, the tsunami ran up to at least 30 m above sea level at the southwest end of Unimak Island [R. Black, written communication, 1980]. This island is directly adjacent to the main portion of the rupture. The observation from Unimak Island clearly establishes a tsunami amplitude at least 3 times larger near the main source than the tsunami amplitude about 200 km away.

A decrease of amplitude of a factor of about 2 over a distance of 200 km seems reasonable, since tsunami amplitude decreases as:

$$h \approx h_0 / (d)^{1/2}$$

where h_0 is the amplitude at the source and d is distance from the source [Solov'ev, 1965]. Tsunami run-up associated with the 1968 Tokachi-oki earthquake decayed by a factor of about 2 over a distance of about 200 km [Kajiura et al., 1968].

We note, also, that local bathymetry can strongly influence tsunami amplitudes [e.g. Hatori, et al., 1973]. Therefore, observations of near source tsunami heights alone do not allow us to distinguish whether tsunami excitation occurred in the Unalaska segment. Our only conclusion in this regard is that near source observations do not contradict the previous result that the Unalaska segment did not excite significant tsunami energy.

Comparison with Nankaido Earthquake of 1946

The coincidence of the easternmost extent of the 1957 tsunami source area and the main portion of the aftershock zone is remarkable and could lead to the premature conclusion that the 1957 earthquake did not rupture the Unalaska segment of the aftershock zone. It is instructive, therefore, to compare the observations of the 1957 earthquake with similar observations from an unusual and well-studied earthquake in Japan, the 1946 Nankaido earthquake. One month after the Nankaido main shock, aftershocks covered an area with a length of 210 km along strike [Kanamori, 1972b; Ando, 1975], whereas geodetic data suggest a rupture length of about 320 km [Fitch and Scholz, 1971]. Of this extended zone, the western 90 km lacked aftershocks, but probably slipped more (by about a factor of 2) than did the rest of the rupture zone [Fitch and Scholz, 1971]. Figure 5 illustrates the region near the Nankaido event and the main features of the earthquake.

Ando [1975] studied the records from two tide gauges located near the aftershock zone of the Nankaido earthquake. One, located near the northern edge of the aftershock zone, shows a simple motion at the start of the tsunami. The other, located at Hosojima, about 250 km southwest of the aftershock zone, shows a more complex

onset. Ando identified two distinct arrivals within the initial portion of the tsunami (see Figure 5). The main energy of the tsunami, arriving at 'b' in the record, was generated at the location labelled 'b', which coincides with the outline of the aftershock area. An earlier, smaller and less distinct arrival that Ando identifies at 'a' seems to originate from a source that extends somewhat to the west of the geodetically determined rupture zone. In addition to lacking aftershocks, the westernmost portion (90 km in length) of the rupture zone was inefficient in generating a tsunami. The absence of a strong tsunami source in the western portion strongly suggests this area ruptured with a very long time constant of at least several minutes in duration [Kanamori, 1972a; 1972b]. A rupture with such a long time constant could involve either a very slow rupture propagation combined with rapid displacement [Fukao, 1979; Das and Scholz, 1981], or a fast rupture propagation with very slow displacements on the fault, possibly as in an episode of creep [Kanamori, 1972b]. The identification of a small tsunami phase excited by the westernmost portion of the rupture zone suggests, however, that the rupture propagated rapidly (compared to tsunami speeds) across the western portion of the zone but because of slow strain release was inefficient at generating a tsunami. Otherwise, the small tsunami excited by the western area would have been concealed by the later and much larger main tsunami generated by the eastern region.

In the lower corner of Figure 5 we compare two California tide gauge records from the 1957 tsunami with that of Hosojima from the 1946 Nankaido tsunami. A precursor comparable to 'a' produced by the 200 km long Unalaska segment would arrive approximately 20 minutes before the main tsunami phase. There is no obvious arrival prior to the main tsunami in the two records of 1957. We conclude that any such phase, if present, is very small and that if any tsunami was generated at all by the Unalaska segment, generation was inefficient.

Discussion

The behavior of the Unalaska segment at the time of the 1957 earthquake seems to have been distinctly different from that of the rest of the aftershock zone. An unusual type of rupture in the Unalaska segment, either as a delayed event or as a very slow source process, might explain the observations. Alternatively, the lack of rupture within the Unalaska segment could, also.

Consider first that the Unalaska segment ruptured in an unusual manner in 1957. The possibility that it broke in a delayed seismic event with a normal-type rupture seems unlikely since it requires a fortuitous complexity of rupture that is not supported by the lack of aftershock activity within the Unalaska segment. Timing of the Unalaska rupture would be critical. Rupture would have to be delayed by at least 25

minutes in order not to interfere with the initial tsunami energy. Rupture would also have to occur quickly enough after the main shock that the seismic and tsunami energy that resulted would be lost in the coda of the main shock. In addition, the smaller tsunami observed at Scotch Cap compared to that at Unimak Island is inconsistent with this delayed normal rupture interpretation.

If the Unalaska segment did rupture in 1957 it seems more likely that it did so in a very slow and smooth rupture process that would leave stress heterogeneities too small to produce aftershocks. The result of such a rupture would be a slow earthquake [Kanamori and Stewart, 1979] whose seismological effects might not be distinguishable from the coda of the main shock. In this interpretation, however, the occurrence of aftershocks at the northern and deeper edge of the plate interface is puzzling. If material at shallower depths undergoes non brittle deformation, why would the deeper material, which is subjected to higher temperature and pressure, undergo brittle deformation? Since the Unalaska area seems to be a tectonic transition zone, perhaps coupling of the plates is poor at shallow depths, but is strong enough along the deeper edge of the interface to rupture seismically.

The alternative possibility, that no rupture occurred within the Unalaska segment, could explain the aftershocks at 40-60 km depths as activity along the deeper edge of a strongly coupled segment of the plate interface. This activity may have been triggered by rupture of the main segment. The smaller magnitude events that occurred to the east of the Unalaska segment (and within the aftershock zone of the 1946 Aleutian earthquake) could be explained in a similar manner. Since most of the plate interface did not rupture, there would be no aftershocks within the segment. Also, activity would not be stimulated in the adjacent trench region. Finally, the precursory swarm near the eastern edge of the main portion of the aftershock zone could have resulted from preparatory failure at this end of the eventual rupture zone. Such failure may also have occurred near the western end of the aftershock zone.

Few data are available to quantify the behavior of this Unalaska segment during the 1957 event. Long and ultra-long period seismic or strain records might help to resolve if it ruptured in a slow event. No suitable geodetic data are known to the authors from this area. A tide gauge operated at the town of Unalaska, near the eastern side of Unalaska Island, showed somewhat less coseismic subsidence than did a tide gauge operated on Adak Island, near the epicenter of the main shock (12 cm compared to 18) [Wahr and Wyss, 1980]. It is not clear, however, that the amount of subsidence at Unalaska could distinguish whether rupture continued into the Unalaska segment, since Unalaska is so near the western boundary of the segment. Data from several temporary tide gauges that were operated for short

periods of time both before and after the 1957 earthquake may help to resolve whether the Unalaska segment ruptured. At least one gauge was located near the northeastern edge of the segment and presumably would not be influenced by slip on the main portion of the 1957 rupture zone [R. Stein, personal communication, 1980].

It is difficult to identify historic earthquakes that definitely ruptured into the Unalaska segment as opposed to earthquakes that may have ruptured only the main segment of the 1957 aftershock zone or the Shumagin Gap. Nevertheless, it is clear that great earthquakes have broken adjacent portions of the Alaska-Aleutian plate margin on both sides of the Unalaska segment. Therefore it would seem unwise to wait until finding definitive evidence that the Unalaska segment has ruptured in a great earthquake before considering that this area is capable of producing one.

The question of whether the Unalaska segment ruptured in 1957 has serious implications for the evaluation of seismic and tsunami hazards in the eastern Aleutians. If it did not break then, it could not have ruptured more recently than 1902, 78 years ago [Sykes et al., 1981]. In this case, not only would the Unalaska segment be a seismic gap, but it could also fall into a category of high potential for producing a large or great earthquake, as described by McCann et al. [1979]. Davies et al. [1981] found a range of repeat times of about 50 to 90 years for great earthquakes within 400 km to the east of the possible Unalaska Gap. So, from a recurrence time consideration it would seem that if it is a seismic gap it may also be a mature seismic gap.

We conclude that at the time of the earthquake of 1957, the 200 km long Unalaska segment behaved in a manner quite different from the rest of the 1200 km long aftershock zone. Although the behavior of this segment may not be resolvable with available data, one of two scenarios seems likely. The Unalaska segment may have ruptured in a slow event in 1957, and as a result would now have a low seismic potential [McCann et al., 1979]. On the other hand, if the segment did not rupture in 1957, it would be a seismic gap with a high seismic potential and hence, could pose significant seismic and tsunami hazards for the eastern Aleutians in the near future.

The dimensions of the unbroken arc segment would be about 210 km in length along the plate margin and 90 km in down dip width. If a single earthquake broke the entire area of the Unalaska segment and had an average stress drop ($\Delta\sigma$) of 20 bars, the displacement across the fault surface would be about 4 m, the moment, M_0 , about 3×10^{28} dyne cm, and the magnitude, M_w , about 8.3 [Kanamori and Anderson, 1975; Kanamori, 1977].

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